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DEGASSING HISTORY AND EVOLUTION OF VOLCANIC ACTIVITIES OF TERRESTRIAL PLANETS BASED ON RADIOGENIC NOBLE GAS DEGASSING MODELS

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ABSTRACT

The radiogenic noble gas species ^4He, ^{40}Ar and ^{129}Xe in the atmospheres of the terrestrial planets provide valuable information on planetary degassing history. The abundance of radiogenic fraction of Martian atmospheric ^{129}Xe relative to the total planetary mass suggests that early degassing, possibly during magma ocean cooling, would be 1/3 as much as that of the Earth. Atmospheric ^{40}Ar amounts relative to the total mass of each planet show that the long-term volcanic degassing of Mars and Venus are 1/20 and 1/4 of that of the Earth, respectively. The duration of plate tectonics is estimated from the total ^{40}Ar amounts: for Mars it is less than 300Ma assuming a plate velocity of 8 cm/s and ridge length of 8000 km and for Venus it is less than 1Ga. Detection of Martian ^4He will be evidence of current degassing activity on Mars, whereas the observed abundance of Venusian ^4He is the outcome of long-term degassing from the interior.

INTRODUCTION

The current atmospheres of the terrestrial planets were formed by secondary degassing from planetary interiors [1,2]. The degassing history of terrestrial planets is divided into two categories. One is early degassing during or within a short period after planetary formation. Impact degassing [3] and/or outgassing during magma ocean cooling are plausible processes for early degassing. The other is continuous (or episodic) degassing throughout planetary history, i.e., degassing in the course of magmatic activity at hot spots and during production of oceanic crust at mid oceanic ridges [4].

The principle volatiles that are degassed from the interior are H_2O and CO_2. Because of their high reactivity and retentivity by surface rocks, H_2O and CO_2 can be trapped and incorporated into solids, forming hydrate or carbonate minerals in the crust; their atmospheric abundances do not reflect the total degassed amount. In the Earth, moreover, H_2O and CO_2 may be recycled or "regassed" into the mantle through plate subduction. In the Venusian atmosphere, H_2O has decomposed and hydrogen escaped into space.

On the other hand, atmospheric noble gases provide important information for planetary degassing history in spite of their low concentrations. Because of their low reactivity, noble gases cannot be trapped by solids or incorporated into the interior again. The present noble gas abundances in the atmosphere reflect degassing activities integrated over the planetary history. As for the lightest rare gas He, continuous escape to the space has decreased the atmospheric abundance. which therefore reflects ongoing present degassing activity.

Among noble gas isotopes, radiogenic species such as ^4He (produced through decay of ^{235}U, ^{238}U, and ^{232}Th) , ^{40}Ar (a decay product of ^{40}K) and ^{129}Xe (a decay product of ^{129}I) can provide chronological information for degassing since their abundances increase with time. Because of the short half life of ^{129}I (1.57×10^7yr), ^{129}Xe can be used in discussing early degassing from the interior [5]. The difference between atmospheric and interior $^{129}Xe/^{130}Xe$ favors the early (<< a few 10^8yr) catastrophic degassing of the Earth, which was originally proposed on the basis of the difference of atmospheric and mantle $^{40}Ar/^{36}Ar$ [6]. Comparing radiogenic ^{129}Xe abundances in their atmosphere, Sasaki [7] concluded that early Martian degassing is about one third as much as that of the Earth.

The atmospheric abundance of ^{40}Ar is a useful clue in pursuing the rate of long-term global volcanic activity because of the longer half life of ^{40}K (1.25×10^9yr). In discussing the total abundance, the physical process of volatile ^{40}Ar transport from the interior to the surface must be examined precisely. Then we may deduce meaningful information on volcanic activity history. In the Earth, most of degassing is associated with volcanic activity forming oceanic crust at mid oceanic ridges. Tajika and Matsui [8] estimated the long-term history of terrestrial magmatic production, taking into account melt generation through adiabatic decompression of uprising mantle materials, partitioning of ^{40}Ar and ^{40}K between silicate melt and solid, and bubbling and degassing of volatiles during magma eruption. They numerically examined changes in mantle temperature, heat flux, melt generation and degassing rates using a simple convection model. They showed that the average magma generation rate over the Earth's age is about 37 km^3/yr, which is a little larger than the current production rate of oceanic crust of 20 km^3/yr. In Mars, on the basis of absolute ages estimated from surface crater density and volume of hot-spot type volcanics, the magma production rate through time is obtained [9]. Sasaki [7] compared the erupted volcanic volume with the present ^{40}Ar abundance in the Martian atmosphere, considering that partial melting should concentrate volatiles. He concluded that the present ^{40}Ar abundance is compatible with the present apparent volume of volcanics, assuming some amount of intrusive volcanics.

In this study, we present quantitative estimates of Martian and Venusian volcanic activities assuming that atmospheric ^{40}Ar abundances constrain the integrated magmatic production. In the Earth, plate-tectonic activity (degassing at mid oceanic ridges should supply more volatiles as well as magma than hot spot type activity; long-term degassing of ^{40}Ar is mainly controlled by plate-tectonic activity [8]. If there were plate-tectonic activity on Mars or Venus, it would have supplied a significant amount of ^{40}Ar into the atmosphere. From the present ^{40}Ar abundance in the atmosphere, we can obtain the upper limit on the duration of plate-tectonic activity on Mars or Venus. Recently, Sleep [10] proposed that the ancient plate tectonics occurred on the northern hemisphere of Mars. Using some of the parameters in his model, we try to constrain duration of the Martian plate tectonics from ^{40}Ar abundance. A similar estimate is briefly shown for Venus, too. Finally we discuss another radiogenic noble gas, ^4He, which constrains current degassing on Mars and possibly long-term degassing on Venus.

Table 1 Normalized abundances of representative noble gas isotopes in CI chondrites, Earth, Venus and Mars: R(X) [kilogram/kilogram-planet] [11].

	^{4}He	^{20}Ne	^{36}Ar	^{84}Kr	^{130}Xe
CI	-	2.89×10^{-10}	1.25×10^{-9}	3.57×10^{-11}	7.0×10^{-12}
		±0.77	±0.10	±0.15	±1.9
Earth	6.21×10^{-10}	1.00×10^{-11}	3.45×10^{-11}	1.66×10^{-12}	1.40×10^{-14}
	±0.07	±0.01	±0.01	±0.02	±0.02
Venus	1.1×10^{-10}	2.9×10^{-10}	2.51×10^{-9}	4.7×10^{-12}	8.9×10^{-14}
	$+2/-0.5$	±1.3	±0.97	$+0.6/-3.4$	$+2.5/-6.8$
Mars	?	4.38×10^{-14}	2.16×10^{-13}	1.76×10^{-14}	2.08×10^{-16}
		±0.74	±0.55	±0.28	±0.41

Table 2 Isotopic ratios of Ar and Xe in the atmospheres of the terrestrial planets

	$^{40}Ar/^{36}Ar$	$^{129}Xe/^{130}Xe$	$^{129}Xe/^{132}Xe$
Earth	295.5	6.50	0.983
Venus	1.19 ± 0.07	-	-
Mars Viking	3.01×10^{3}	-	1.5-4.5
EETA79001	2.26×10^{3}	16.40 ± 0.8	2.39 ± 0.03

RADIOGENIC ^{129}Xe AND ^{40}Ar ABUNDANCES AND DEGREE OF DEGASSING

Table 1 summarizes noble gas abundances of CI chondrites, Earth, Mars and Venus [11]. Each abundance is expressed as the mass of species X in the atmosphere divided by the total planetary mass (R(X): abundance of X in kilogram/kilogram-planet). Table 2 shows the isotopic ratios involving ^{129}Xe and ^{40}Ar. For Venus, we have no data on Xe isotopes. Although almost all of ^{40}Ar is radiogenic, there is a significant primary non-radiogenic component of ^{129}Xe. According to Pepin and Phinney [12], only 6.7% of ^{129}Xe in the terrestrial atmosphere is the decay product of ^{129}I. Assuming that the initial non-radiogenic component is the same between Mars and Earth ($^{129}Xe^{n}/^{130}Xe = 6.06$, where superscript n denotes nonradiogenic species) as a first approximation, we obtain the abundance of radiogenic ^{129}Xe (with

$^{129}Xe^*/^{130}Xe$ = 10.34, where superscript * denotes radiogenic species) of the Martian atmosphere from the total abundance of Martian ^{129}Xe (with $^{129}Xe/^{130}Xe$ = 16.40) [7]. A similar analysis is made also by Pepin [13]. Then 63% of ^{129}Xe in the Martian atmosphere is of radiogenic origin and we have

$$\frac{R\left({}^{129}\mathrm{Xe}^*\right)_{Mars}}{R\left({}^{129}\mathrm{Xe}^*\right)_{Earth}} = \frac{R\left({}^{129}\mathrm{Xe}\right)_{Mars} \times 0.63}{R\left({}^{129}\mathrm{Xe}\right)_{Earth} \times 0.067} = 0.353 \qquad (1)$$

A similar value was obtained by Zahnle [14]. In contrast to the isotopic ratio, the Martian radiogenic ^{129}Xe abundance is less than the terrestrial value. But the Martian atmosphere still contains one third of radiogenic ^{129}Xe compared with the terrestrial atmosphere in relative abundance (kg/kg-planet).

Because we may consider that all of ^{40}Ar is the radiogenic product of ^{40}K, the relative ratio of ^{40}Ar is more easily obtained both for Mars and Venus: we have

$$\frac{R\left({}^{40}\mathrm{Ar}^*\right)_{Mars}}{R\left({}^{40}\mathrm{Ar}^*\right)_{Earth}} = 0.048 \qquad \text{and} \qquad \frac{R\left({}^{40}\mathrm{Ar}^*\right)_{Venus}}{R\left({}^{40}\mathrm{Ar}^*\right)_{Earth}} = 0.29 \qquad (2)$$

In the above we use data of SNC meteorite EETA79001 for Martian Ar. The Martian atmosphere is deficient not only in non-radiogenic Ar but also in radiogenic Ar; it contains only one twentieth of ^{40}Ar in relative abundance (kg/kg-planet) compared with the terrestrial atmosphere [15-17]. This ^{40}Ar deficiency can be ascribed to either less tectonic activity on Mars or atmospheric escape. Although the Venusian atmosphere contains 70 times as much ^{36}Ar as than the terrestrial atmosphere, it has one fourth of ^{40}Ar compared with the terrestrial atmosphere.

Radiogenic noble gas isotopes are supplied into the atmosphere through degassing from the interior. The above difference between relative Martian ^{129}Xe and ^{40}Ar abundances suggests that atmospheric ^{129}Xe was not supplied into the atmosphere by the inefficient degassing process that has supplied ^{40}Ar. Using a simple two-stage degassing model where sudden degassing occurred at $t = t_1$ and $t = t_2$ ($> t_1$), we try to express the above ratios (1) and (2) by parameters describing the degassing process [7]. Here α is the degassing fraction from the interior at the initial degassing at $t = t_1$, and β is that at the later degassing at $t = t_2$ ($> t_1$), where t_1 is longer than the half life of ^{129}I and much shorter than the half life of ^{40}K and t_2 is at least comparable to the half life of ^{40}K. We assume that the degassing fractions (α and β) are the same for argon and xenon: the fraction α (β) of gas species (radiogenic or non-radiogenic) degassed suddenly at $t = t_1$ ($t = t_2$). So long as the orders of magnitude of α and β are not largely different, we have for the present abundances of radiogenic ^{129}Xe and ^{40}Ar in the Martian atmosphere:

$$\begin{aligned} {}^{129}\mathrm{Xe}^* &= \alpha\, {}^{129}\mathrm{I}_0\left(1-\exp(-\lambda_{129}t_1)\right) \\ &\quad + \beta\left[(1-\alpha)^{129}\mathrm{I}_0\left(1-\exp(-\lambda_{129}t_1)\right) + {}^{129}\mathrm{I}_0\left(\exp(-\lambda_{129}t_1)-\exp(-\lambda_{129}t_2)\right)\right] \qquad (3) \\ &\approx \alpha\, {}^{129}\mathrm{I}_0\left(1-\exp(-\lambda_{129}t_1)\right) \end{aligned}$$

and

$$\begin{aligned} {}^{40}\mathrm{Ar}^{*} &= \alpha f_{\mathrm{Ar}}{}^{40}\mathrm{K}_0\left(1-\exp(-\lambda_{40}t_1)\right) \\ &+ \beta\left[(1-\alpha)f_{\mathrm{Ar}}{}^{40}\mathrm{K}_0\left(1-\exp(-\lambda_{40}t_1)\right)+f_{\mathrm{Ar}}{}^{40}\mathrm{K}_0\left(\exp(-\lambda_{40}t_1)-\exp(-\lambda_{40}t_2)\right)\right] \\ &\approx \beta f_{\mathrm{Ar}}{}^{40}\mathrm{K}_0\left(\exp(-\lambda_{40}t_1)-\exp(-\lambda_{40}t_2)\right) \\ &\approx \beta f_{\mathrm{Ar}}{}^{40}\mathrm{K}_0\left(1-\exp(-\lambda_{40}t_2)\right) \end{aligned} \tag{4}$$

where λ_{129} and λ_{40} are the decay constants and ${}^{129}I_0$ and ${}^{40}K_0$ are the initial abundances of ${}^{129}I$ and ${}^{40}K$, respectively, and $f_{Ar} = 0.1048$ is the branching ratio for ${}^{40}Ar$. Assuming that the abundances of ${}^{129}I_0$ and ${}^{40}K_0$ relative to the total planetary mass are the same among the Earth, Mars and Venus, we have

$$\frac{\mathrm{R}\left({}^{129}\mathrm{Xe}^{*}\right)_{Mars}}{\mathrm{R}\left({}^{129}\mathrm{Xe}^{*}\right)_{Earth}} = 0.353 \approx \frac{\alpha_{Mars}}{\alpha_{Earth}}$$

$$\frac{\mathrm{R}\left({}^{40}\mathrm{Ar}^{*}\right)_{Mars}}{\mathrm{R}\left({}^{40}\mathrm{Ar}^{*}\right)_{Earth}} = 0.048 \approx \frac{\beta_{Mars}}{\beta_{Earth}} \quad \text{and} \quad \frac{\mathrm{R}\left({}^{40}\mathrm{Ar}^{*}\right)_{Venus}}{\mathrm{R}\left({}^{40}\mathrm{Ar}^{*}\right)_{Earth}} = 0.29 \approx \frac{\beta_{Venus}}{\beta_{Earth}} \tag{5}$$

In the above, the timings of degassing events among the planets are assumed to be the same ($t_{M1} \sim t_{E1} \sim t_{V1}$ and $t_{M2} \sim t_{E2} \sim t_{V2}$). So long as $t_{M1}, t_{E1}, t_{V1} \sim 10^8$yr and $t_{M2}, t_{E2}, t_{V2} \geq 2\times10^9$yr, the differences among t_{M1}, t_{E1}, t_{V1} and among t_{M2}, t_{E2}, t_{V2} do not change the above relations. From the above derivations, it is confirmed that differences in the current abundances of radiogenic noble gases among the Earth, Mars, and Venus should correspond to differences in the degassing fraction (α or β) from the interior. And the earlier and later degassing event can be discussed separately using relative abundance ratios of ${}^{129}Xe$ and ${}^{40}Ar$.

The ratio of radiogenic ${}^{129}Xe$ between the Earth and Mars (0.353) suggests that the outgassed fraction at the early degassing ($t = t_1 \sim 10^8$yr) on Mars is less than that of the Earth but the difference is not so large. Impact degassing during accretion and degassing during magma ocean cooling are plausible mechanisms for the early degassing. Since impacts would not only supply volatiles but remove atmospheric gases of the smaller planet Mars by "atmospheric cratering" [14, 18-20], impacts should produce larger differences in atmospheric volatile inventories between the Earth and Mars. On Mars, atmospheric cratering would have decreased non-radiogenic noble gas abundances greatly and may explain their present small abundances [14]. In this respect, we prefer the magma ocean cooling after the accretion - atmospheric cratering stage as the mechanism of early degassing, which supplied radiogenic ${}^{129}Xe$ into the atmosphere. Because accretion timescale ($\sim10^7$yr) is comparable to the half life of ${}^{129}I$, the atmospheric cratering should have also escaped some radiogenic ${}^{129}Xe$; this loss might account for a part of the difference of relative radiogenic ${}^{129}Xe$ abundances between the Earth and Mars (denoted by the relation (5)). Another idea is that I/Xe fractionation occurred on Mars by the difference of water solubility between iodine and xenon [15, 21]: whereas primary xenon was lost from the atmosphere by the atmospheric cratering, iodine should have dissolved into the protoocean, been captured by crustal rocks, and finally supplied radiogenic ^{129}Xe. There is no data on Venusian ^{129}Xe, nor is precise data of Xe elemental abundance. Therefore we cannot discuss about the early degassing on Venus using radiogenic ^{129}Xe abundance.

We assumed that the initial mass abundances (kg/kg-planet) of ^{129}I were the same between the Earth and Mars. From relatively higher iodine abundance of SNC meteorites than terrestrial mantle, Dreibus and Wänke [15] estimated mean abundance of Martian iodine 2.4 times higher than that of the Earth. This might decrease the corresponding ratio of degassing degree $\alpha_{Mars}/\alpha_{Earth}$ in the relation (5) proportionally. Moreover, the initial ^{129}I is affected by the formation age, i.e., the time when iodine is trapped firmly into solids in the cosmochemical stage. We assumed that this age difference would be small in the scenario of the solar system formation. The case where the radiogenic ^{129}Xe difference is ascribed to the different formation ages was discussed previously [7].

Different from radiogenic ^{129}Xe, the above relation (5) suggests that later degassing on Mars deduced from ^{40}Ar abundances is much less than that of the Earth. This corresponds to a smaller volume of melting on Mars after the initial hot epoch during or just after accretion. Indeed, assuming that the present ^{40}K abundance is the same as that of the Earth (5.0×10^{-8} kg/kg), the present Martian atmospheric ^{40}Ar abundance ($R_{Mars}(^{40}Ar) = 5.7\times10^{-10}$ kg/kg) is only about 1% of the total radiogenic ^{40}Ar produced in the interior of 5.0×10^{-8} kg/kg. Later degassing on Venus is also smaller than that of the Earth, and corresponds to a current absence of plate tectonics. In the following sections, using more detailed models, we discuss the history of volcanic activities on Mars and Venus and place upper limits on the duration of their plate-tectonic epoch, if they existed.

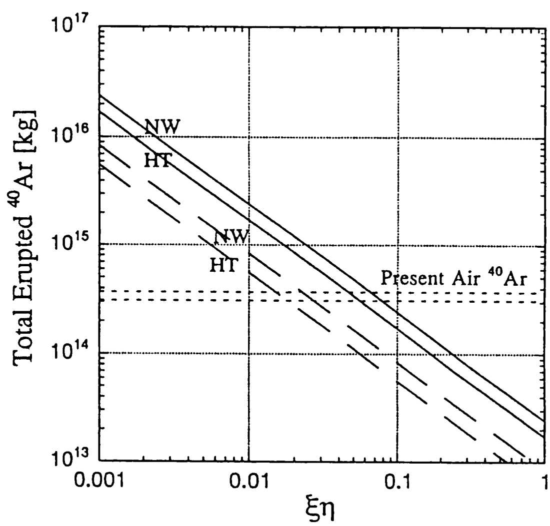

Fig.1. The total degassed ^{40}Ar for different values of $\xi\times\eta$. Solid lines are based on the volume estimate by Greeley [9]. The dashed lines are based on the volume estimate by Greeley and Schneid [22]. NW denotes the age model by Neukum and Wise [23] and HT denotes the age model by Hartmann et al. [24] and Tanaka [25].

MARTIAN ^{40}Ar AND VOLCANIC ACTIVITY

From surface images of Mars, the erupted volcanic volume at each epoch has been estimated [9, 22]. According to Greeley [9], the total erupted volume of volcanics is 2×10^{17} m^3. Combined with an age model based on crater density, we can estimate the magma generation rate associated with Martian volcanoes, which are apparently hot-spot type. Sasaki [7] compared the total erupted volcanic volume with the present ^{40}Ar abundance in the Martian atmosphere. The outgassed ^{40}Ar mass by volcanic activity is estimated using the product of two parameters, the melt accumulation factor (~ melt fraction) ξ and the contribution of intrusive volcanics (ratio of the extrusive mass to the total erupted mass) η (Fig. 1). When the volcanic eruption rate at the surface is denoted by q, the total magma supply (generation) rate is q/η. The mantle volume, which supplies magma during unit time, is expressed by $q/\eta\xi$, and then ^{40}Ar degassing rate is approximated by $C_{Ar}q/\eta\xi$, where C_{Ar} is ^{40}Ar concentration. Then, the degassing ^{40}Ar amount is inversely proportional to $\xi\times\eta$ (see Fig. 1). In Fig. 1, the initial ^{40}K abundance of Mars is assumed to be the same as that of the Earth. Assuming η ~ 1/4 - 1/2, a favorable value for melt fraction of ξ ~ 0.1 - 0.2 could explain the present Martian ^{40}Ar atmospheric abundance. Scambos and Jakosky [17] compared the erupted volcanics with the present ^{40}Ar abundance in order to constrain non-radiogenic volatiles such as ^{36}Ar and H_2O. Although they obtained a parameter "^{40}Ar release factor" which corresponds to $\xi\times\eta$ in Sasaki [7], they did not pursue discussion on the history of Martian volcanic activity. Pepin [13] also used the erupted volcanic volume of Mars to obtain the degassed ^{40}Ar from the interior.

In the above estimate, we cannot assume an arbitrary small value for η because there is little evidence of intrusive volcanic volumes on Mars. Previously very small value (~ 1/9.5) is used by Greeley and Schneid [22], but it is not possible that Martian large volcanoes should involve nearly large intrusive volumes. A part of discrepancy between the erupted volume and degassed ^{40}Ar might be explained using a larger value for the initial ^{40}K abundance in Mars. However, there is a possibility of other magmatic sources for ^{40}Ar. Recently, Sleep (1994) proposed that a large part of the Martian northern hemisphere had been formed by plate-tectonic activity during the earliest history of Mars; the area is about one third of the total surface of Mars. Like the Earth, ridge magmatism involving crustal formation should supply much volatiles into the atmosphere. Subducted materials may account for the assumed intrusive volcanics (denoted by η) in the previous model.

Here we consider that the present atmospheric ^{40}Ar is supplied both by identified large volcanoes (hot spots) and by ancient plate tectonics. Then we can obtain an upper limit of the duration of the ancient plate-tectonic activity. First, the amount of ^{40}Ar degassed by hot-spot type volcanism is evaluated from the erupted volcanic volume [9], using the cratering age model [24, 25]. It is expressed as

$$M_{Ar-h} = \int_0^{4.6\mathrm{Ga}} F_{Ar-h}\,dt$$

$$F_{Ar-h}(t) = fC_{Ar}(t)k_h \approx C_{Ar}(t)q_h(t)\xi_h^{-1} \qquad (6)$$

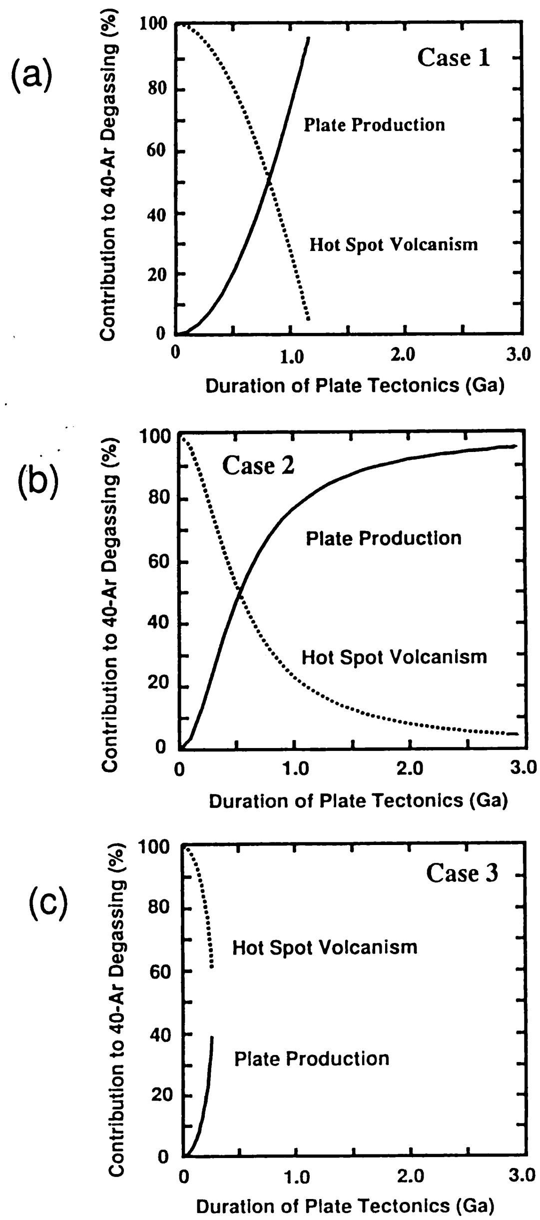

Fig. 2 Contribution of two types of volcanism (hot spot type and plate tectonics type) to ^{40}Ar degassing as a function of the duration of plate tectonics on Mars. (a), (b), and (c) correspond to the results for Cases 1, 2, and 3 respectively.

where subscript h denotes hot spot, $F_{Ar\text{-}h}(t)$ is the ^{40}Ar degassing rate, $C_{Ar}(t)$ [kg/m^3] is the ^{40}Ar concentration in the mantle, ξ_h is the accumulation factor (~ melting degree), k_h [m^3/yr] is the volume rate of uprising mantle material that actually contributes to magma production, and then $q_h(t) = \xi_h k_h$ [m^3/yr] is the melt production rate. The efficiency of ^{40}Ar degassing from the uprising volume, f, is close to the unity [8] and is neglected here. The amount of outgassed ^{40}Ar is inversely proportional to ξ_h.

The degassing rate of ^{40}Ar by plate-tectonic activity is written similarly as

$$M_{Ar-p} = \int_0^{\tau} F_{Ar-p}\, dt$$

$$F_{Ar-p}(t) = f\, C_{Ar}(t) k_p \approx C_{Ar}(t) q_p(t) \xi_p^{-1} \quad (7)$$

where the subscript p denotes plate tectonics and τ represents the duration of plate tectonics. In estimating supply rate of magma at Martian ridges, we assume three cases for the volume rate k_p (Tajika and Sasaki, in preparation).

In the first case, we use the Earth's production rate $q = 2\times10^{10}$ m^3/yr for Mars and the melt fraction $\xi = 0.2$ of oceanic crust (assuming a higher mantle temperature of the early Mars), together with a surface area correction:

$$\text{Case 1} \qquad k_p = 2.0\times10^{10}\,[1/3(R_M/R_E)^2](0.2)^{-1}\ [\text{m}^3/\text{yr}] \quad (8)$$

where a ratio of planetary radii of Mars and the Earth, R_M/R_E, is 0.53, and the area of plate tectonics is restricted to the northern lowlands (~1/3 of the surface) of Mars [10].

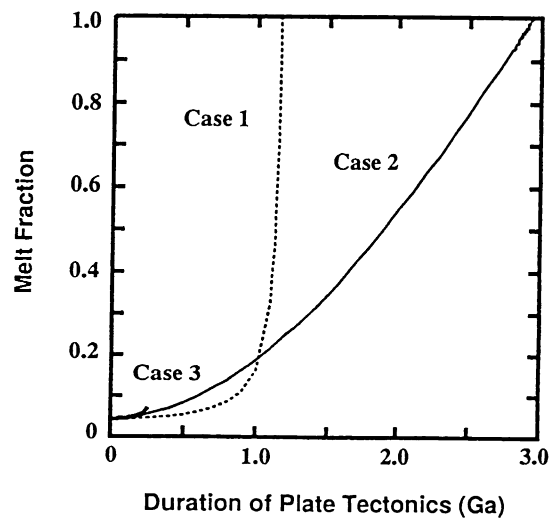

Fig. 3 Summary of the average melt fraction of Martian volcanism as a function of the duration of plate tectonics for Cases 1, 2, and 3. It is noted that in Case 3 the solution is absent over some critical duration of about 0.3Ga.

In the second case, we assume that the Earth's production rate applies to Mars but use the same melt fraction ξ as that of hot spot type activity. Then ξ is a variable parameter also for the plate-tectonic supply.

$$\text{Case 2} \qquad k_p = 2.0\times10^{10}\,[1/3(R_M/R_E)^2]\xi_h^{-1}\,[\mathrm{m^3/yr}] \qquad (9)$$

In Cases 1 and 2, the melt fraction ξ is an independent variable. But the application of the terrestrial production rate is too rough an assumption. Next we use a model which combines geological parameters [10] and detailed model of the melt generation process beneath oceanic ridges [26]. We use an average plate velocity v_p = 8 cm/yr and the length of oceanic ridges of 8000 km [10]. Using the interior solidus curve calculated for Martian low gravity, we can estimate the generated melt thickness d as well as the melt fraction ξ (= the generated melt thickness / the melt generation depth) from the assigned mantle potential temperature T_p, which is the temperature of mantle material expanded adiabatically to the surface pressure. The calculation is based on the model proposed by McKenzie and Bickle [26] (Tajika and Sasaki, in preparation). We have

$$\text{Case 3} \qquad k_p = 0.08\times8\times10^6\,d(T_p)\,\xi(T_p)^{-1}\,[\mathrm{m^3/yr}] \qquad (10)$$

where d and ξ are function of mantle potential temperature T_p. In the above, the mantle potential temperature T_p is an independent variable and $\xi(T_p)$ should also affect the hot spot type eruption rate k_h. For all numerical calculations, time step is taken to be 1.0 Ma.

Our results are shown in Figs. 2 and 3. Since the total degassed ^{40}Ar amount (3.6×10^{14} kg) is fixed, variations of ξ or T_p will change contributions of hot spot and plate tectonics to ^{40}Ar degassing and change the duration of plate tectonics (Figs. 2). The corresponding melt fraction ξ is shown in Fig. 3. In Case 1 (Fig. 2(a)), the hot spot production rate k_h decreases with increasing ξ while the plate production rate k_p is constant. The duration of plate tectonics τ is as long as 1Ga

Table 3 Critical (maximum) plate tectonics duration τ_{cr} and corresponding melting degree ξ_{cr} in Case 3

		τ_{cr}	ξ_{cr}
1	Standard Mars	270Ma	0.086
2	Earth's solidus	440Ma	0.085
3	$[^{40}Ar] = 2\times$present $[^{40}Ar]$	560Ma	0.044
4	$[^{40}Ar] = 0.5\times$present $[^{40}Ar]$	125Ma	0.168
5	v_p = 0.8 cm/yr	880Ma	0.084

for $\xi_h \sim 0.2$ (see Fig.3). In Case 2 (Fig. 2(b)), both k_h and k_p decrease with increasing ξ. The duration of plate tectonics would be much longer but plausible ξ_h (≤ 0.2) should keep τ not longer than 1Ga (see Fig.3). In Case 3 (Fig. 2(c)), k_h should decrease with increasing T_p through ξ but k_p should increase with increasing T_p through enhancing melt thickness. Increasing k_p results in a relatively short plate tectonics duration τ (< 0.3Ga) and a critical value for ξ ($\xi_{cr} \sim 0.07$) (see Fig.3). As seen in Fig. 3, no solution exists above the critical value ($\xi > \xi_{cr}$ and $\tau > \tau_{cr}$). Assuming that the start of plate-tectonic activity is at $t = 0$ ($C_{Ar}(0) = 0$), we obtain the longest duration of Martian plate tectonics. If we assume a later time for the beginning of plate tectonics ($t > 0$), then plate activity would outgas more ^{40}Ar and its duration would be shorter.

We obtain plate tectonics duration τ for different constraining parameters in Case 3. The results are shown in Table 3 where the critical duration τ_{cr} is tabulated. Under the terrestrial solidus curve, a smaller volatile accumulation volume results in a smaller degassing rate; we have a longer duration. We have examined this case for different values of the total abundance of atmospheric ^{40}Ar. Jakosky et al. [27] propose that pick-up-ion sputtering by solar winds and photochemical escape should decrease atmospheric species. Using isotopic fractionation of $^{38}Ar/^{36}Ar$, about half of ^{40}Ar has been lost from the Martian atmosphere (Jakosky, personal communication). On the other hand, if the initial degassing event was intense ($\alpha >> 0.1$) and prolonged, early degassed ^{40}Ar would have occupied several tens of percent of the total atmospheric ^{40}Ar, which would decrease the constraining amount of later degassed ^{40}Ar. Then we simulated the cases when the total ^{40}Ar is twice and half as large as the present amount (see Table 3). The former prolongs the plate tectonics duration τ (~0.6Ga) and the latter shortens τ (~0.1Ga). Finally, we make sure that a slower assumed plate velocity would prolong the duration τ greatly, suppressing the ^{40}Ar degassing rate.

VENUSIAN ^{40}Ar AND VOLCANIC ACTIVITY

It is more difficult to estimate the erupted magmatic volume of Venus because global resurfacing has eliminated any previous record of volcanic activity. Tajika and Matsui [28] assumed that Venusian tectonic history can be divided into two stages: Stage I when plate tectonics prevails and Stage II when volcanic activity is limited to the hot spot type. During Stage I ($0 < t < \tau$) with plate tectonics, we apply the terrestrial magma production rate. According to the study using ^{40}Ar degassing model coupled to the thermal history of the Earth, the seafloor spreading rate averaged over 4.6Ga is constrained to be close to the present value [8], although this result might not guarantee the application of the terrestrial production rate for Venus. During State II ($\tau < t < 4.6\times10^9$yr), we use the ^{40}Ar degassing rate $F_{Ar}(t) = C_{Ar}(t)k_h(t)$ (Eq.(6)). Temporal variations of k_h and ξ are estimated from the thermal history calculation; ξ is variable here for the model of Venus although it was constant for the model of Mars. Because the total outgassed ^{40}Ar amount is fixed and equal to the present atmospheric amount, the duration of plate-tectonic activity constrains the average k_h and then the magma eruption rate q_h.

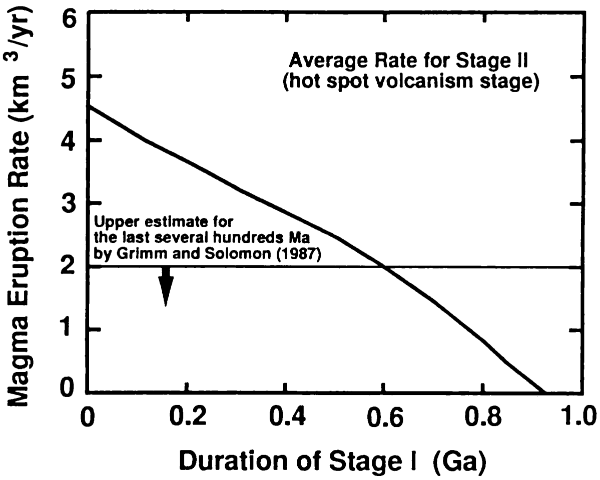

Fig. 4. The average magma eruption rate of Venus estimated from the ^{40}Ar degassing model as a function of the duration of Stage I (the presumed plate tectonics stage on Venus). The solid curve represents the magma eruption rate averaged over the Stage II (the stage after plate tectonics ceased). The horizontal line represents the upper bound of melt production activity estimated from crater statistics on Venus by Grimm and Solomon [29].

Our result is shown in Fig. 4. The horizontal axis shows the duration of plate tectonics and the vertical axis corresponds to the average magma generation rate over Stage II. The duration of ancient plate tectonics τ was less than 1Ga for Venus if the ridge spreading rate was the same as that of the Earth over the same time period. If the duration of Stage I, τ, exceeds 1.0Ga, almost all of the ^{40}Ar degassed into the atmosphere during that stage, suggesting that volcanic activity after $t = \tau$ should be nearly "zero" (note that a longer duration is obtained if we assume the lower magma production rate for Stage I). Longer plate-tectonic activity is incompatible with a young surface age of Venus, which suggests recent global volcanic activity. If plate tectonics did not operate on Venus, the magma generation rate averaged over Stage II is estimated to be about 5 km^3/yr. Longer plate-tectonic activity, required for degassing the observed amount of ^{40}Ar, results in an average magma generation rate smaller than 5 km^3/yr. Grimm and Solomon [29] considered the Venusian younger surface age inferred from crater statistics to be due to volcanic resurfacing and estimated the recent melt generation rate on Venus to be smaller than 2 km^3/yr. If this limit is applicable to the whole of Stage II, some duration of plate-tectonic activity (Stage I) associated with efficient ^{40}Ar degassing is necessary.

MARTIAN AND VENUSIAN ^4He

The mass spectrometer on board Viking lander did not detect ^4He in the Martian atmosphere [30]. Recently, the EUVE (Extreme Ultraviolet Explorer) satellite observed He airglow (58.4nm resonance scattering) of Mars and detected

43±10R (1R = 10^6 photons/cm^2 s) [31]. This corresponds to a ^4He abundance of 1.1±0.4ppm, so that the present Martian atmosphere contains 2.8×10^{10} kg of ^4He.

In contrast to the heavier ^{129}Xe and ^{40}Ar isotopes, atmospheric ^4He, which is more easily lost to space, does not reflect long-term degassing history. In the Earth, ^4He escapes along with magnetic field lines from polar regions as "polar wind" [32] and ^4He outflow was observed by polar orbiters [33]. The life time of ^4He in the atmosphere is estimated to be 10^6yr [34]. Mid oceanic ridges supply about 20-30% of ^4He whereas a large part of atmospheric ^4He is considered to come from the continental crust where U and Th are enriched [34-36]. Although mid oceanic ridge volcanism can supply ^4He continuously over the timescale of 10^6yr, some mechanism such as magmatic transport or water circulation is necessary to maintain continuous degassing from the continental crust because of the slow ^4He diffusion rate [34].

Since a dipole magnetic field is absent on Mars, ^4He cannot be lost by polar winds. Because of low gravity, however, electron impact ionization followed by solar wind pick-up can cause the escape of the atmospheric ^4He on a timescale of 5×10^4yr [31]. Then the detection of ^4He implies that the current degassing rate is continuous in this timescale. Since at present Mars does not have an apparent continuous degassing source such as mid oceanic ridges, the existence of Martian ^4He requires some other mechanism of supply. We estimated degassing from impact cratering. Using a crater density model and assuming that ^4He escapes from volumes ten times as large as that of craters, we have a degassing rate of 10-30 kg/yr. Then the ^4He supply timescale is $(1\text{-}3)\times10^9$yr, which is much larger than the estimated lifetime of 5×10^4yr. We speculate that ongoing underground magmatism or hydrothermal circulation is necessary to account for the ^4He abundance.

We need more accurate measurements of Martian He. In the EUVE observations, the size of Mars was less than one pixel and there was a high noise level. Future fine mass spectrometric detection by a lander or UV observation by an orbiter will confirm the existence of Martian ^4He, which should constrain the current degassing from the interior.

In Venus, the atmospheric ^4He abundance is much higher than the Earth: we have $R_{Venus}(^4He)/R_{Earth}(^4He) = 200$. Because this ratio is even larger than that of Ne or Ar, a large part of atmospheric ^4He is not primary and degassed later from the interior. Because of the absence of a magnetic field, the ^4He escape rate from the current Venus atmosphere should be much slower than that of the Earth; Prather and McElroy [37] estimated the escape rate and obtained 6×10^8yr for the life time of ^4He in the Venusian atmosphere. The lifetime can be also estimated from the degassing rate. If the degassing rate of ^4He is compatible with that of ^{40}Ar and if ^4He inventory is in steady state in the Earth's and Venusian atmospheres, the life time of atmospheric ^4He in Venus $t_{He\text{-}V}$ can be approximately estimated from the terrestrial life time $t_{He\text{-}E}$ which is better constrained: taking $t_{He\text{-}E} \sim 10^6$yr, we have $t_{He\text{-}V} \sim [R_{Venus}(^4He)/R_{Venus}(^{40}Ar)] \times[R_{Earth}(^{40}Ar)/R_{Earth}(^4He)] \times t_{He\text{-}E} \sim [200 / 0.29] \times10^6 \sim 7\times10^8$yr. Since this value, which is compatible with the previous estimate, is shorter than 4.6×10^9yr, the present atmospheric ^4He abundance in Venus can be explained by a radiogenic degassed component.

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